Factors Controlling the Depths and Sizes of Magma Reservoirs in Martian Volcanoes. L. Wilson[1,2] and J.W. Head[2]. [1]Environmental Science Division, Institute of Environmental and Biological Sciences, Lancaster University, Lancaster LA1 4YQ, U.K. L.Wilson@Lancaster.ac.uk [2]Department of Geological Sciences, Brown University, Providence RI 02912, U.S.A. James_Head@Brown.edu

We investigate how the variation of density with depth controls the locations of magma reservoirs in basaltic volcanoes on Mars, finding that reservoir centers should typically lie at depths of 9 to 13 km, and discuss stress controls on the vertical extents of reservoirs.

Magma reservoirs in basaltic volcanoes on Earth commonly develop with their centers at depths at which magmas are neutrally buoyant in the surrounding crust [1]. This implies that hydrostatic stresses dominate dynamic stresses in determining where in the crust the dikes and sills which ultimately develop into reservoirs come to rest, and there are theoretical reasons related to the ease with which dike tips propagate why this is likely to be the case [2]. A simple model [3] of the variation of crustal density ρ with depth h in basaltic volcanic areas assumes that the volcanic pile grows by deposition of fragmental or vesicular eruption products which then compact under gravity so that the void space V decreases exponentially with pressure P: $V = V_0 \exp(-\lambda P)$ where V_0 is the surface void space fraction and λ is a constant. If ρ_∞ is the limiting density of the crustal rocks at infinite compression, the definition of void space implies that the surface density $\rho_{surf} = \rho_\infty (1-V_0)$, and in general $\rho = \rho_\infty(1-V)$. Suitable integration of the definition of the pressure variation, $dP/dh = \rho g$, leads to

$$\rho(h) = \rho_\infty/[1 + \{V_0/(1-V_0)\} \exp(-\lambda\rho_\infty gh)]$$

$$P(h) = (1/\lambda) \ln[V_0 + (1-V_0) \exp(\lambda\rho_\infty gh)]$$

and the best fit [3] of these functions to density profiles in Hawai'i and Iceland on Earth, for which ρ_{surf} ~2200 kg m^{-3} (so that $V_0 = 0{\cdot}24$) and ρ_∞ ~2900 kg m^{-3}, gives $\lambda = 11{\cdot}8$ GPa^{-1}. We expect that λ will have a similar value on Mars, since the potential for density-altering processes other than simple compaction (e.g. hydrothermal fluids and secondary mineral deposition) existed for the martian volcanoes.

Table 1 shows the subaerial crustal density variations implied by this function using the observed $V_0 = 0{\cdot}24$ for Earth. A rising basaltic magma with a typical density of 2600 kg m^{-3} will be neutrally buoyant at a depth of 3 km, and it is at just this depth that the magma reservoir in the hawaiian volcano Kilauea is centred [4]. The corresponding density profile given for Mars (where the lower atmospheric pressure may lead to relatively vesicular eruption products [5]) uses a value of 0.325 for V_0. Also shown are the bulk densities as a function of depth for two candidate martian magmas which have liquid densities of 2600 kg m^{-3} (making them buoyant in the 2900 kg m^{-3} density crust at depth) and dissolved CO_2 contents in their source zone of 0.1 and 0.3 wt%. The symbols P, N and - indicate whether the magma is positively, negatively or neutrally buoyant in the crust. Progressive exsolution of the volatile with decreasing pressure (exacerbated by assuming that the volatile is the relatively insoluble CO_2) reduces the bulk magma density and, at the higher volatile content, causes the magma to reach a neutral buoyancy level (NBL) shallower than the 11.16 km depth appropriate to gas-free magma. Unless the magma is very CO_2-rich, however, magma reservoir growth is likely to be centred close to the 11 km NBL for this surface void fraction. Reducing V_0 decreases the NBL, to ~8 km for $V_0 = 0.25$; similarly increasing V_0 to 0.5 increases the NBL to ~16 km. These values should represent extremes of the range for Mars, suggesting that most magma reservoirs should be centred at depths between perhaps 9 and 13 km.

Magma reservoirs grow incrementally as a result of injections of fresh magma from below which increase the internal fluid pressure to the point where new dikes propagate from the boundary. As a new dike cools it becomes part of the country rocks, but the thermal gradient near the boundary of the reservoir ensures that no dike cools to the point where no trace of its existence remains. The remains of previous dikes represent the best sites for subsequent dike growth because their tips, although becoming ever more rounded as cooling progresses, are locations of stress concentration at the reservoir margin. Driving pressures (country rock stress minus magma pressure) of only a few MPa are needed to re-activate meter-scale dike stubs [6], and repeated growth of dikes at a single point on the reservoir wall in this way leads to lateral rift zone formation. If old dike stubs cool beyond the point where they represent favored locations for wall rock failure, the reservoir walls must fail in tension or compression, requiring much larger stress differences across the boundary and making the site of failure (controlled by the stresses due to edifice topography) less predictable. This implies that martian sheild volcanoes without well-defined rift zones are those which are most likely to have had long repose periods between episodes of recharge and growth.

Geometric factors, plus the fact that the magma inside a reservoir tends to be slightly less compressible than the country rocks of the growing volcanic pile, means [7] that a smaller stress difference is needed to cause failure around the middle of a reservoir than at its roof or floor. In the lowest-stress-difference case of dike stub reactivation, the absolute pressure inside a martian reservoir centred at a depth of 11 km must be close to the country rock stress level, ~97 MPa. The equivalent for a reservoir on Earth centred at 3 km depth is 72 MPa. These pressures consist of the weight of the magma within the upper half of the reservoir plus an excess pressure in the magmatic fluid. On Earth, where hawaiian reservoirs have half heights of ~2 km, the magma weight is ~52 MPa implying an excess pressure of ~20 MPa. This excess pressure must be inherited from the processes of melt formation by pressure-release partial melting in the mantle. If similar processes, and hence excess pressures, occur on Mars, this implies magma weights of ~77MPa. This in turn implies that reservoir half-heights could be as large as 7 to 8 km, placing reservoir roofs as shallow as 3 to 4 km.

These geometric considerations have implications for the pressure gradients driving magma to the surface and into lateral dike systems, and hence for eruption rates. We anticipate using these results as guidelines in interpreting the locations, sizes and geometries of volcanic deposits.

References. [1] Ryan, M.P. (1987) p. 259 in Magmatic Processes: Physico-Chemical Prin-ciples. [2] Lister, J.R. (1990) JFM 210, 263. [3] Head, J.W. & Wilson, L. (1992) JGR 97, 3877. [4] Rubin, A.M. & Pollard, D.D. (1987) USGS Prof. Pap. 1350, 1449. [5] Wilson, L. & Head, J.W. (1983) Nature 302, 663. [6] Pollard, D.D. (1987) p.5 in Geol. Assoc. Can. Spec. Pap. 34. [7] Parfitt, E.A., Wilson, L. & Head, J.W. (1993) JVGR 55, 1.

Table 1. Crustal density (in kg m^{-3}) as a function of depth (in km) in basaltic volcanic areas on Earth and Mars, calculated using surface rock void fraction values discussed in the text, and densities of basaltic magmas containing 0.1 and 0.3 wt% CO_2 in their source zones. The underscore marks the neutral buoyancy level at which a magma reservoir is likely to be centred.

depth	Earth crust	Mars crust	magma 0.1wt%		magma 0.3wt%	
0	2200	1958	3	P	1	P
1	2364	2037	2442	N	2131	N
3	2600	2184	2571	N	2453	N
5	2739	2313	2597	N	2528	N
8	2840	2472	2600	N	2571	N
10	2868	2558	2600	N	<u>2584</u>	<u>N</u>
11	2877	2594	2600	N	2589	P
11.16	2878	2600	<u>2600</u>	<u>-</u>	2590	P
12	2884	2628	2600	P	2593	P
15	2894	2709	2600	P	2600	P